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A DIAGNOSIS OF THE LIFE CYCLE OF A MARINE CYCLONE SYSTEM

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1. INTRODUCTION

This paper describes a diagnosis, using the Zwack-Okossi methodology, of the life cycle of a cyclone that occurred over the North Atlantic Ocean during October 1985. This case was chosen because it serves as a precursor to the blocking case presented by the current authors earlier in this meeting.

2. DATA AND METHODOLOGY

The data for this study were obtained from 2° lat x 2.5° lon NASA/Goddard Laboratory for Atmospheres analyses (Schubert et al., 1993). Analyzed fields were provided at mandatory levels and then interpolated linear in ln (p) to 50 mb increment isobaric surfaces. The study period extended from 1200 UTC 29 October 1985, 6 h before the appearance of the surface cyclone studied here, through 1800 UTC 31 October, when the cyclone was in a fully decaying state.

The primary diagnostic tool used was the relatively new development equation known as the Zwack-Okossi (Z-O) equation (Zwack and Okossi, 1986; Lupo et al., 1992). As derived in Lupo, et al., the complete Z-O equation is a combination of the vorticity and thermodynamic equations that equates the relative geostrophic vorticity tendency ($\partial\delta_g/\partial t$) at a near-surface pressure level (p_ℓ = 950 mb) to a set of dynamic (represented by the vorticity equation, terms a-f) and thermal (represented by the thermodynamic equation , terms g-i) forcing mechanisms vertically integrated from p_ℓ to an upper pressure level chosen sufficiently high to encompass most of the atmospheric mass (p_t=30 mb):

$$\left.\frac{\partial\delta_g}{\partial t}\right)_{p_\ell} = \mathrm{PD}\int_{p_t}^{p_\ell}\Big[\underbrace{-\bar{V}\cdot\nabla\delta_a}_{\substack{a\\ \text{vadv}}}\underbrace{-\omega\frac{\partial\delta_a}{\partial p}}_{\substack{b\\ \text{vvte}}}\underbrace{+\delta_a\frac{\partial\omega}{\partial p}}_{\substack{c\\ \text{dive}}}$$

$$\underbrace{-\bar{k}\cdot\left(\nabla\omega\wedge\frac{\partial\bar{V}}{\partial p}\right)}_{\substack{d\\ \text{tilt}}}\underbrace{+\bar{k}\cdot(\nabla\wedge\bar{F})}_{\substack{e\\ \text{fric}}}\underbrace{-\frac{\partial\delta_{ag}}{\partial t}}_{\substack{f\\ \text{ageo}}}\Big]dp \qquad (1)$$

$$-\frac{(\mathrm{PD})R}{f}\int_{p_t}^{p_\ell}\left[\int_{p}^{p_\ell}\nabla^2\left(\underbrace{-\bar{V}\cdot\nabla T}_{\substack{g\\ \text{tadv}}}\underbrace{+\frac{\dot{Q}}{c_p}}_{\substack{h\\ \text{diab}}}\underbrace{+S\omega}_{\substack{i\\ \text{adia}}}\right)\frac{dp}{p}\right]dp$$

PD is $1/(p_\ell - p_t)$, $\bar{V}$ the horizontal wind vector, ω the vertical motion (dp/dt), δ_a the absolute vorticity, δ_{ag} the ageostrophic vorticity, $\bar{F}$ the frictional force, T the temperature, $\dot{Q}$ the diabatic heating, S the static stability parameter ($-T/\theta\ \partial\theta/\partial p$), θ the potential temperature, ∇ the del operator on an isobaric surface, R the dry air gas constant, c_p the specific heat at constant pressure, and f the coriolis parameter.

In addition, the geostrophic vorticity tendency at p_i = 500 mb was diagnosed from the equation

$$\left.\frac{\partial\delta_g}{\partial t}\right)_{p_i} = \left.\frac{\partial\delta_g}{\partial t}\right)_{p_\ell}$$

$$+\frac{R}{f}\int_{p_i}^{p_\ell}\nabla^2\left(-\bar{V}\cdot\nabla T+\frac{\dot{Q}}{c_p}+S\omega\right)\frac{dp}{p}. \qquad (2)$$

Finally, both (1) and (2) were relaxed to produce solutions for the height tendencies at p_ℓ and p_i.

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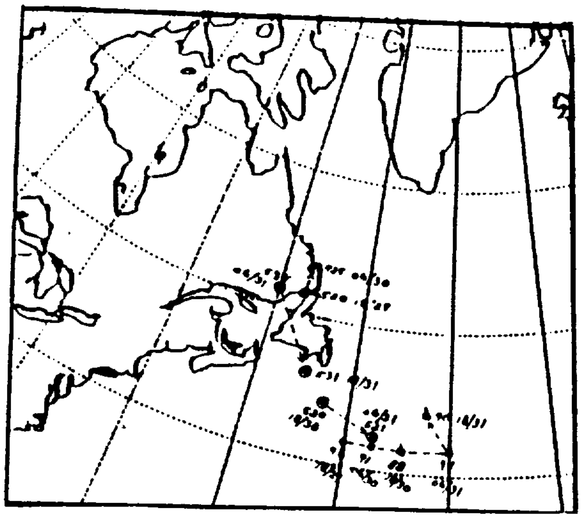

Fig. 1. Surface (•) and 500 mb (⊗) low positions with central sea-level pressures (mb) and 500 mb heights (dm) in 12 h increments from 1800 UTC 29 October (18/29) through 1800 UTC 31 October (18/31) 1985.

In (1) and (2), ω was calculated using a complete form of the omega equation, while $\vec{F}$ was restricted to boundary-layer processes and calculated using the scheme proposed by Krishnamurti (1968). Diabatic heating (diab)included convective and grid-scale latent heat release (lath, see Lupo et al., 1992), boundary-layer sensible heating (sens, see Lupo, et al.), and infrared radiation (irch, based on the Sasamori (1968) parameterization with the Harshvardhan, et al. (1987) cloud random overlap scheme). All horizontal (vertical) derivatives were calculated with fourth (second)-order finite differencing and vertical integrals were obtained using the trapezoidal rule.

Finally, of particular interest is the relative contribution of each vertical level to surface development, thus documenting the "upper-air support" for the surface cyclone. This is easily seen in terms 1a-f by vertically profiling the integrand. However, this is less easily accomplished with terms 1g-i because of the double integration. To isolate the individual level contributions for these terms, the trapezoidal rule was used to represent the two integrations as the finite sum:

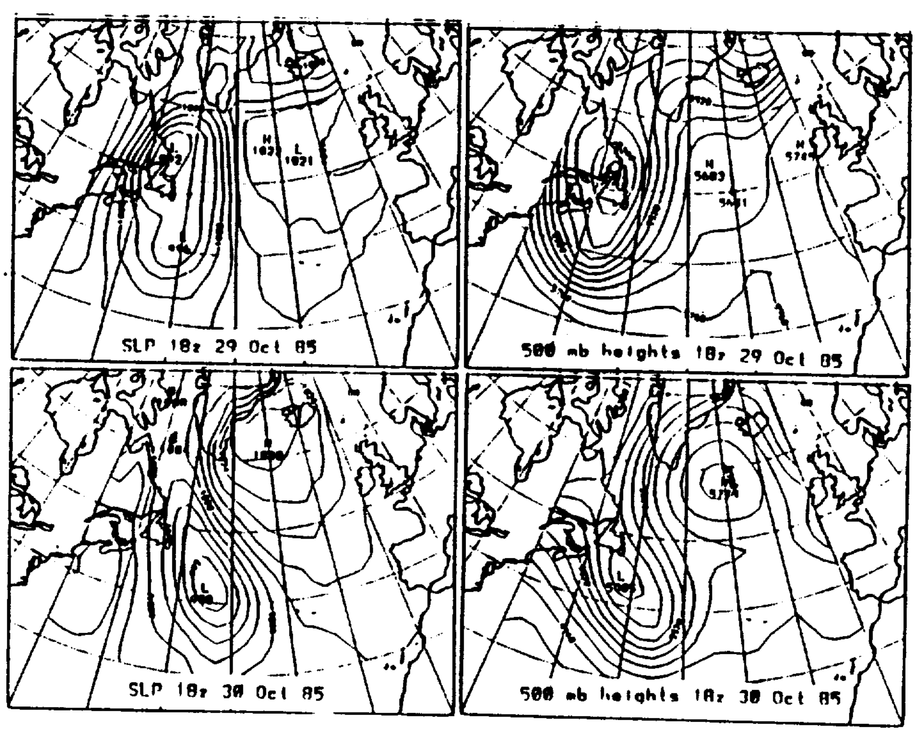

Fig. 2. Sea-level pressure (mb) and 500 mb height (m) analyses for 1800 UTC 29 and 30 October 1985.

$$\frac{R}{f}(PD)\nabla^2\left[(2N-3/4)\frac{H_\ell}{p_\ell}+\sum_{n=2}^{N-1}(N-n)\frac{H_n}{p_n}\right.$$

$$\left.+\frac{1}{4}\frac{H_t}{p_t}\right](\Delta p)^2 \qquad (3)$$

where H is a heating quantity, Δp is 50 mb, N is the total number of levels and n identifies each individual level (n=1 corresponds to p_ℓ and n=N corresponds to p_t). Here the bracketed quantity at each level is multiplied by DPΔp in order to provide profiled heating contributions that can be directly compared with profiles of H_n/p_n, the heating contribution of each level represented in the alternate diagnostic approach suggested by Hirschberg and Fritsch (1991).

3. SYNOPTIC DISCUSSION

The surface cyclone of interest first appeared at 1800 UTC 29 October 1985 about 700 km southeast of Newfoundland (42°N, 50°W) with a sea level pressure of 996 mb (see Figs. 1 and 2). Through the next 24 h the cyclone intensified to 988 mb as it moved eastward to 42°N, 45°W. Thereafter, the cyclone's central pressure increased as it moved further eastward and then northward. Note in Fig. 2 an earlier cyclone located northeast of Newfoundland at 1800 UTC 29 October. This cyclone was in a decaying state and virtually disappeared by 0600 UTC 30 October.

At 500 mb at 1800 UTC 29 October a cyclone was located north of Newfoundland coincident with the earlier surface cyclone. This 500 mb low had already achieved minimum height and was filling as it moved northward (see Fig. 1). This filling continued until the low disappeared at 1200 UTC 30 October and a new low formed southeast of Newfoundland (see 1800 UTC 30 October position in Fig. 2). By 0600 UTC 31 October this second low had moved southeastward as it also filled, and a third 500 mb low is seen northwest of Newfoundland. Finally, by 1800 UTC 31 October the second low had disappeared, and the third low deepened as it moved to just south of Newfoundland.

4. RESULTS

In the interest of efficiency, the results will be confined to vertical profiles of the consistently largest forcing terms (Figs. 3 and 4) and bar graphs of the integrated contribution of each of the terms in (1) and (2) (Figs. 5 and 6). In both cases the values are for the center of the surface cyclone.

Fig. 3 shows that cyclone development was encouraged by cyclonic vorticity advection above 600 mb and warm air advection which maximized at 700 and 150 mb (latent heat release was small at this time). At the same time, adiabatic cooling in the ascending air, maximizing at 600 mb, produced height rises. Profiles associated with the cyclone's decay (Fig. 4) are similar to those for cyclone development except for two notable exceptions, the larger latent heat release and the marked cold air advection below 400 mb.

An additional feature of interest in the thermal terms is a comparison between the weighting given these processes in the Z-O equation (see eq. 3, dashed profiles in Figs. 3 and 4) and the weighting that corresponds to the Hirschberg-Fritsch equation (solid profiles in Figs. 3 and 4). Because of the inverse-pressure weighting, the latter places greater emphasis on thermal processes in the upper troposphere/lower stratosphere. However, the multiple counting of the thermal processes implied by the double integration in (1) results in low-level forcing being "counted" more than high-level forcing in the Z-O equation. Thus, the Z-O profiles assign greater relative importance to lower levels.

From Fig. 5 it is clear that the four terms included in Figs. 3 and 4 were consistently the dominant development mechanisms. Furthermore, the positive development influence of vorticity and temperature advection were more than sufficient to overcome the inhibiting influence of adiabatic cooling during cyclone development (1800 UTC 29 October). However, during cyclone decay (1800 UTC 30 October) the cold air advection below 400 mb reversed the sign of the temperature advection term, which when combined with adiabatic cooling resulted in the decay of the cyclone. Other terms, while not always negligible, were consistently smaller.

At 500 mb, vorticity and temperature advection were again consistently dominant, although several other dynamic terms were close behind. The net result was a relatively large height fall on 29 October near the area where the second 500 mb low formed. By 30 October the significantly smaller height fall reflects the slow eastward propagation of the upper wave.

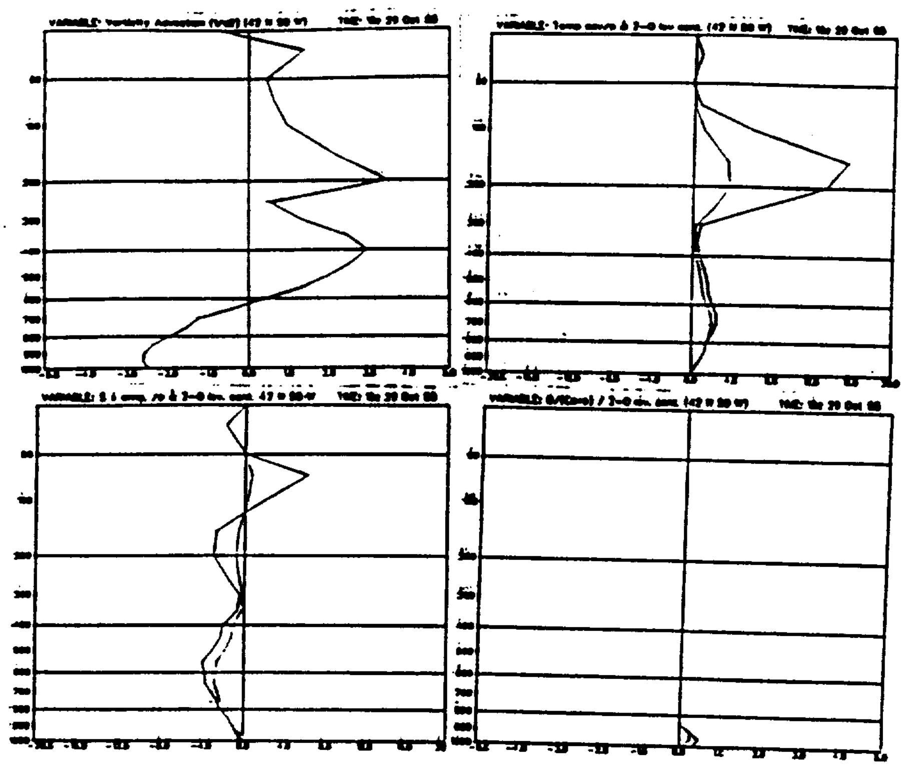

Fig. 3. Vertical profiles of horizontal vorticity advection (upper left, 10^{-9} s^{-2}); inverse pressure-weighted horizontal tamperature advection (upper right, 10^{-7} °K s^{-1} mb^{-1}, Zwack-Okossi form dotted, Hirschberg-Fritsch form solid); adiabatic cooling (lower left, 10^{-7} °K s^{-1} mb^{-1}); and latent heat release (lower right, 10^{-7} °K s^{-1} mb^{-1}) at the cyclone center point for 1800 UTC 29 October 1985. Ordinate is pressure (mb).

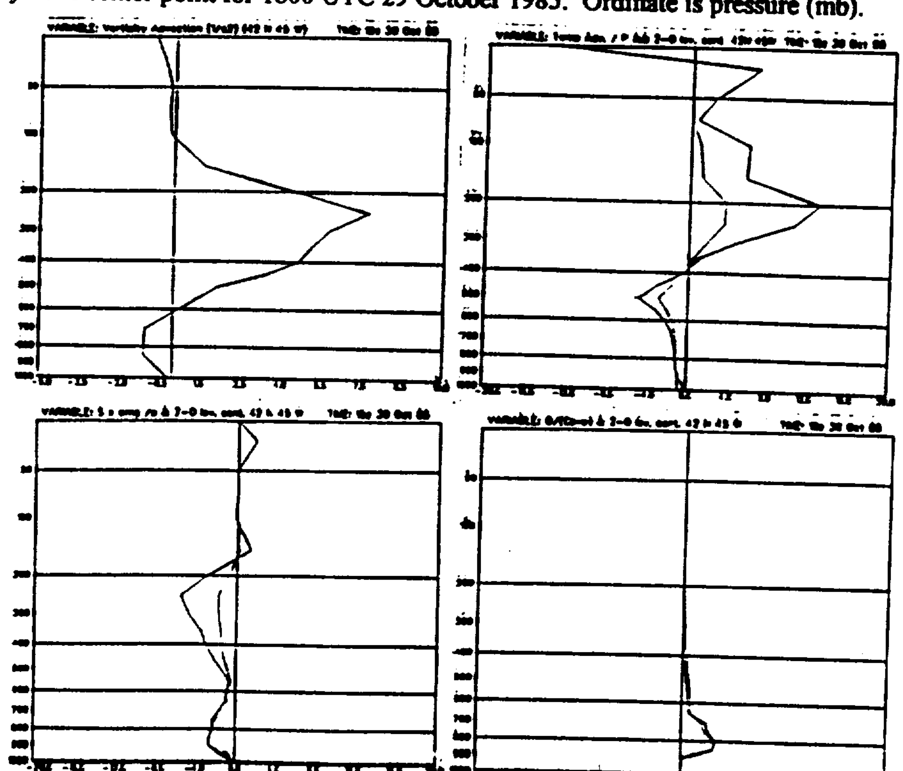

Fig. 4. As in Fig. 3 for 1800 UTC 30 October 1985.

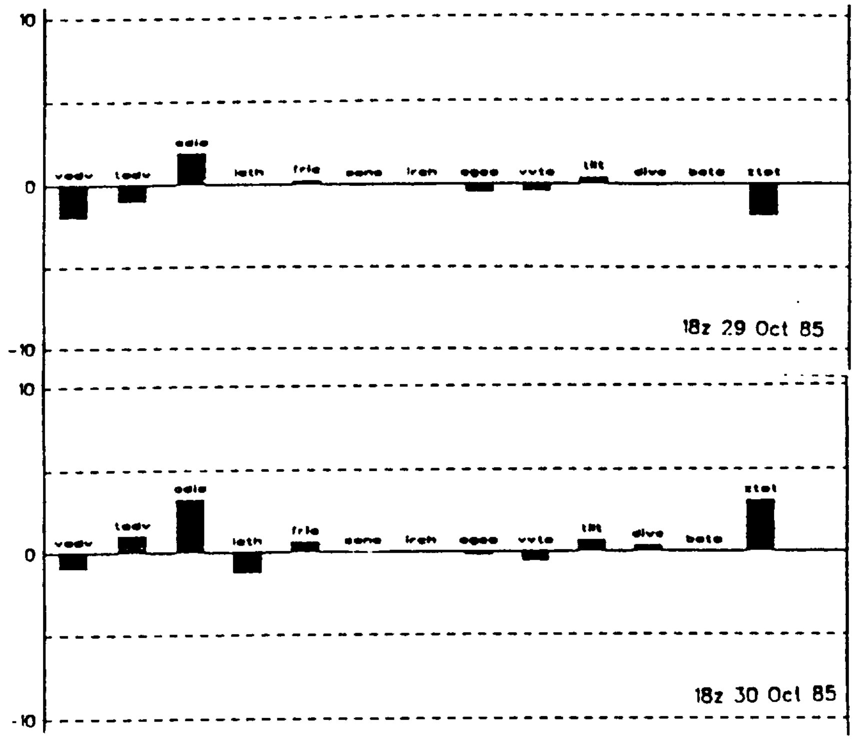

Fig. 5. 950 mb total height tendency (ztot) and contributions from each forcing term in (1) at the cyclone center point. Units: 10^{-3} m s^{-1}.

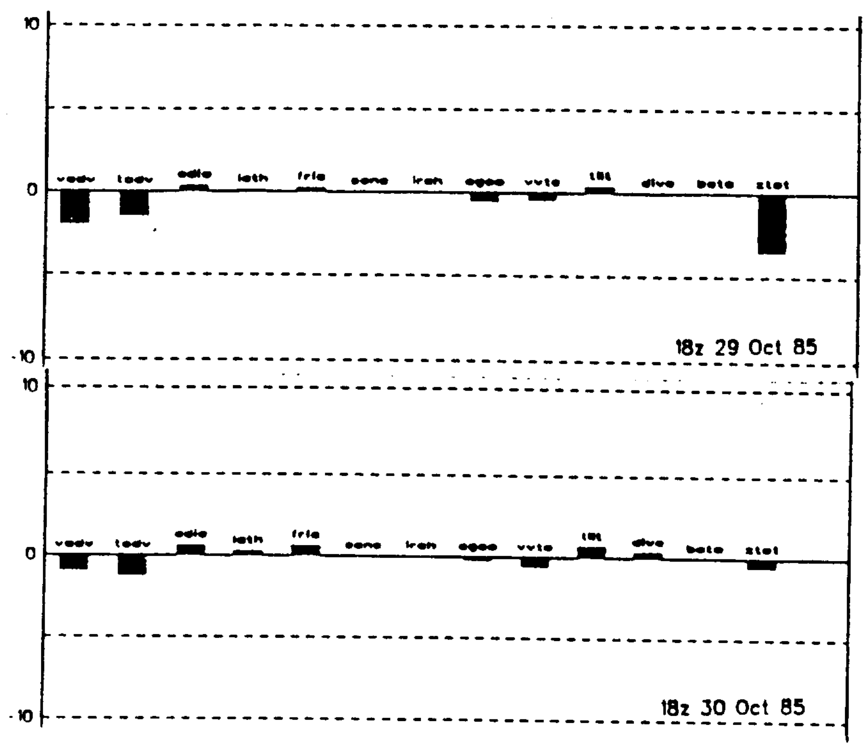

Fig. 6. As in Fig. 5 for 500 mb.

5. CONCLUSIONS

Results confirm previous results of Lupo, et al. (1992) which ascribe cyclone development to vertically-integrated cyclonic vorticity advection, warm air advection, and latent heat release, which combine to overcome the inhibiting influence of adiabatic cooling and other less significant dynamic processes. Then cyclone decay commenced when cold air invaded tropospheric layers below 400 mb.. In addition, while a particular thermal process (e.g., temperature advection) may be greater at high levels, its influence on near-surface height changes may be comparable or in some instances larger at lower levels.

6. ACKNOWLEDGMENTS

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